

# Planet Earth

Don L. Anderson

*"It is my opinion that the Earth is very noble and admirable . . . and if it had contained an immense globe of crystal, wherein nothing had ever changed, I should have esteemed it a wretched lump and of no benefit to the Universe."*

— Galileo Galilei

A PLANET'S SURFACE provides geologists with clues as to what is happening inside. But many of these clues are ambiguous because so many other processes (impacts and erosion, for example) contribute to surface characteristics. Most of the surface of the Earth is less than 100 million years old, and even its oldest rocks are less than 4 billion years old, so the record of the origin of our planet has been erased many times. Part of this erasure is due to erosion by wind and water, and part is due to the continual recycling of material back into the interior and the repaving of the ocean basins by seafloor spreading.

The solar system's other solid planets and smaller worlds have much more ancient surfaces in general, and this tells us two things that we cannot learn from the Earth itself. One is that in the early days of the solar system, violent and destructive impacts were common as larger bodies swept up and devoured smaller ones. The other is that most other worlds preserved evidence of these early happenings, while ours did not.

Geophysicists try to sidestep the mixed signals they encounter at ground level by studying the interiors of planets directly. On our home planet this effort has been rather successful: the Earth is the only body for which we have detailed information, including three-dimensional images of the internal structure, from the surface to the center. And while we realize that other planets do not all share Earth's present behavior, they may at least have been put together in similar ways. So intensive study of the Earth's interior may yield knowledge that is applicable elsewhere, just as our cognizance of the Earth will in turn be shaped by the study of other worlds.

Earlier views of Earth's origin envisioned a gentle rain of dust and small particles that slowly accumulated layer by layer. A planet growing this way would remain relatively cool, building up heat mainly by the slow decay of radioactive elements. According to this scenario, an initially cold, homogeneous Earth eventually heated up, started to

melt, and formed a buoyant crust and a dense core in a way that somehow left behind (in most versions of this story) a homogeneous mantle.

However, planetary evolution has not been so simple. The energy associated with a single large impact is enough to melt, or even vaporize, much of both the impactor and the planet it strikes. If the Moon really came into being when a Mars-size object struck the Earth (see Chapter 4), the energy from that collision would have melted much of the Earth itself. Even smaller hits, of which there were many more, would have caused widespread melting where they penetrated and generated shock waves.

Did the Earth start out cold or hot? The answer depends on whether it accreted slowly (100 million to 1 billion years) or rapidly (100,000 to 10 million years). In the latter case, kinetic energy would have been delivered faster than the growing Earth's ability to conduct and radiate it away as heat, so our planet would have remained molten, at least in its outer parts, as it accreted (Figure 1). In addition, however, every giant impact is essentially an instantaneous accretion event, and planets that grew by gathering up relatively large objects experienced widespread melting over and over again.

We assume, therefore, that growing planets were molten—at least partially, and at least once. At such times, their component materials had the opportunity to separate according to melting points and densities. The "heavy" materials sank toward the interior, creating cores, and the "light" ones rose to the surface, creating crusts. This process of gravitational separation is usually called *differentiation*, and it played a key role in the early histories of the Earth and other terrestrial planets. Over time these worlds may have acquired more internal stratification than simply a light crust and a dense core; as we shall see, the mantle situated between them can itself become layered according to chemistry and density.

Seismology is the geophysicist's principal tool for probing planetary interiors. In a sense, the Earth is a huge spherical bell that is periodically "struck" by earthquakes. We learn about the interior by listening to how the Earth "rings"—that is, by noting how seismic waves move away from the source point, or focus, of an earthquake (Figure 2). Of the four types of seismic waves, two travel around the Earth's surface like the rolling swells on an ocean. A third type, called primary or



*P* waves, alternately compress and dilate the rock or liquid they travel through, just as sound travels. Secondary or *S* waves propagate through rock (but not liquids) by creating a momentary sideways displacement or shear, like the movement along a rope that is flicked at one end. Both *P* and *S* waves slow down when moving through hotter material, and they are refracted or reflected at the boundary between two layers with distinct physical properties.

In fact, we have relied on physical properties such as density and seismic velocity, rather than chemistry or composition, to distinguish the three principal divisions of the Earth's interior. The *crust*, *mantle*, and *core* account for 0.4, 67.1, and 32.5 percent of the planet's mass, respectively. Rocks' physical properties vary with depth due to increasing temperature and pressure and, in places, changes in chemistry or physical state. For example, the most common minerals in the crust and upper mantle are all unstable farther down. As pressure increases the atoms in their crystals become more tightly packed, and their density increases. These changes are gradual except at *phase transitions* (such as when carbon transforms from graphite to diamond under pressure). Phase transitions cause a rapid or abrupt change in physical properties, including those measured by seismic techniques. If the change occurs abruptly, it is called a *seismic discontinuity*.

In the early 1980s, several groups of researchers discovered that seismic waves could be used to produce three-dimensional maps of the Earth's interior, a technique known as *seismic tomography*. The word "tomography" derives from the Greek word for a cutting or section, and in effect geophysicists create a series of cross-sections of the interior at various depths (Figure 3). Seismic tomography is a very powerful technique that has revolutionized our study of the Earth's interior.

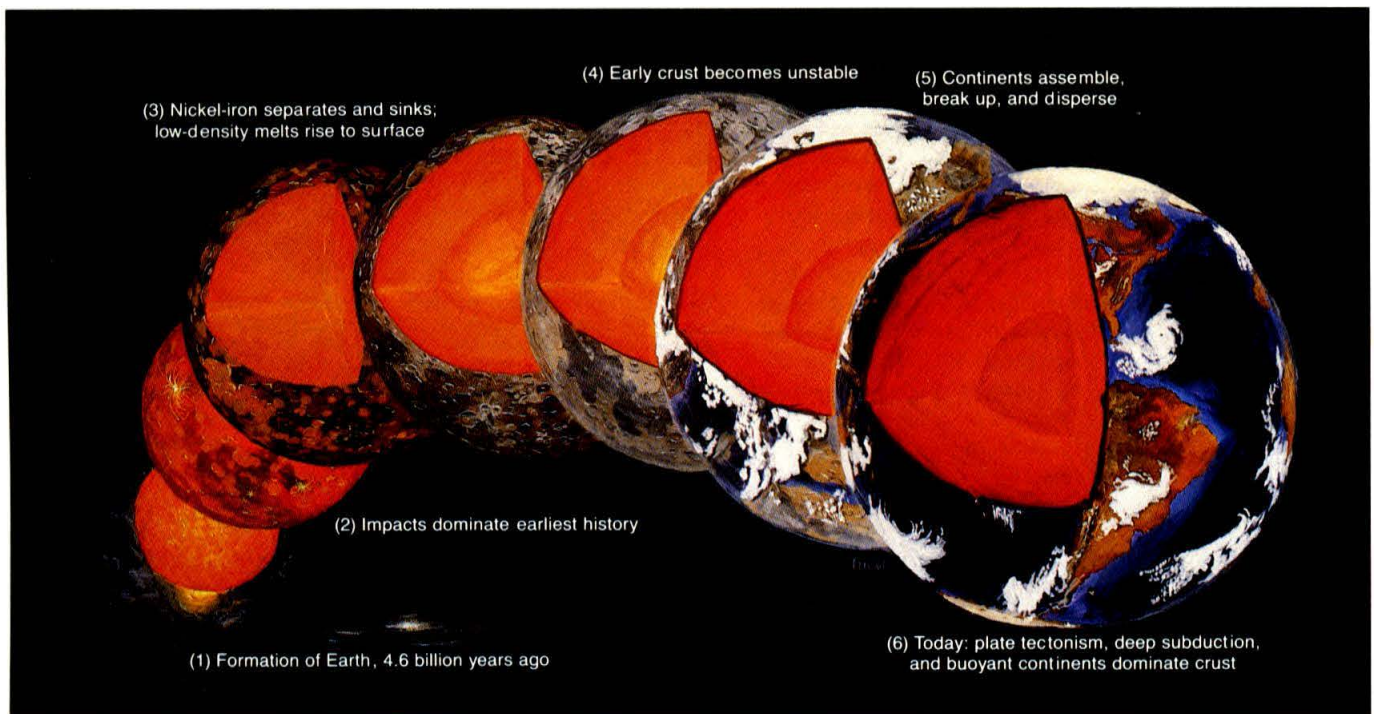
## COMPOSITION OF THE EARTH

Ours is the only planet for which we can speak with some confidence about its bulk composition or chemical makeup. By combining the Earth's mass with seismic determinations of the radius and density of the core, we have deduced that the Earth is about one-third iron and that this iron is concentrated toward the center of the planet. In fact, the Earth's solid inner core, which is smaller in size than the Moon but three times denser, may be pure iron and nickel. Seismology also tells us that the outer part of the core is liquid and indicates strongly that it is molten and mostly iron. To explain its lower density and molten state, the outer core needs to incorporate a small amount of oxide, silicate, or sulfide material.

The Earth is the largest terrestrial planet and contains slightly more than 50 percent of the mass in the inner solar system, excluding the Sun. Compared to Earth, the dense planet Mercury contains proportionately more iron; Mars and the Moon contain substantially less iron, even though they may have small cores. Based on its similarity to our planet in size and density, Venus probably has an Earthlike core. But a solid inner core may be absent, because we expect Venus to have slightly lower pressures and possibly higher temperatures in its interior.

The bulk of the Earth is contained in its mantle, the region between the core and the thin crust. We can sample the top of the mantle in several ways. Fragments of it are exposed in eroded mountain belts and brought to the surface by volcanic eruptions. The major mantle minerals excavated in these ways are olivine  $(\text{Mg,Fe})_2\text{SiO}_4$  and pyroxene  $(\text{Mg,Fe})\text{SiO}_3$ ; thus, iron is present but only as a minor constituent.

The most abundant material we see emerging from the mantle is *basalt*, and it must exist there in vast quantities.



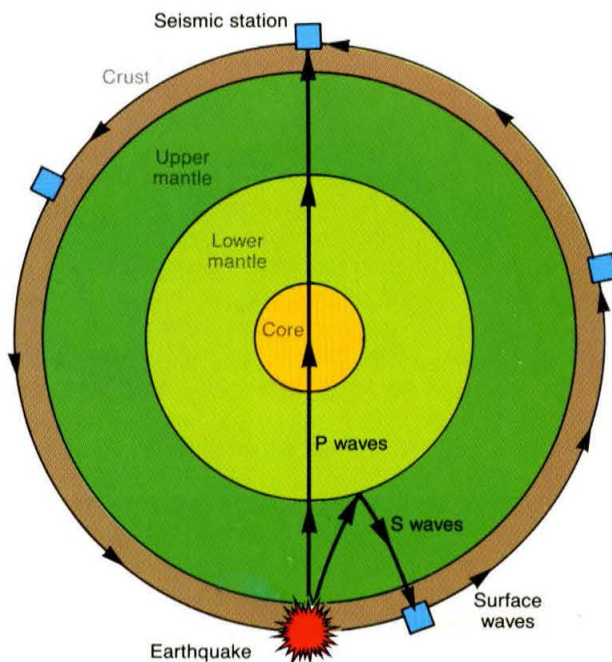
**Figure 1.** Geophysicists do not yet know the exact circumstances of Earth's formation, but our planet's exterior must have been completely molten at least early in its history. Much of the energy needed to melt its outer layers came from innumerable collisions with interplanetary material left over from planetary formation.



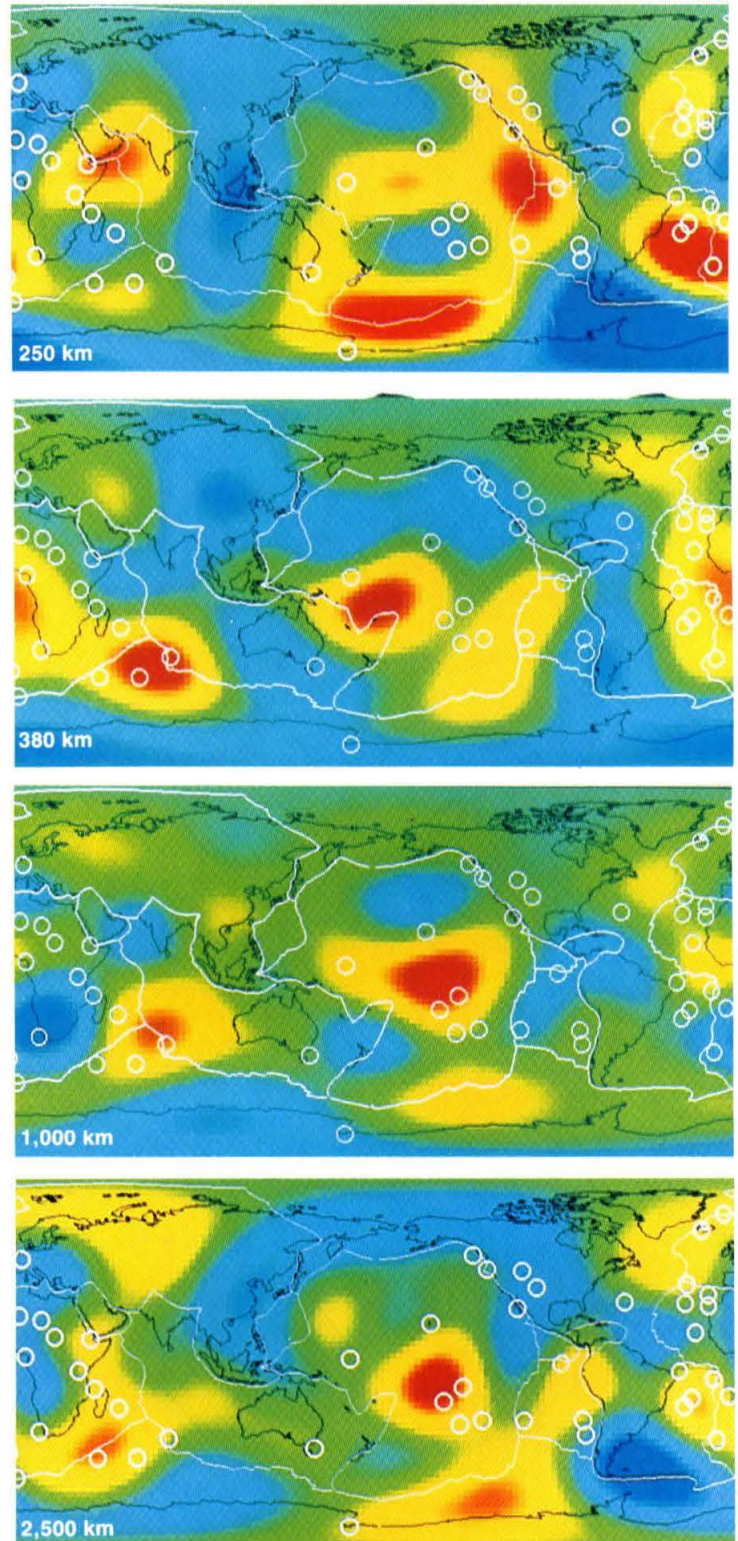
Basaltic magma is rich in the elements calcium and aluminum and is less dense than upper-mantle material, which allows it to erupt into or onto the crust. The ocean floor is covered with basalt. Iceland and Hawaii (Figure 4) are two examples of thick basalt piles that have accumulated on the ocean floor. Hidden from our view under seawater is a 40,000-km-long network of volcanoes – the oceanic ridge system – which generates new oceanic crust at the rate of  $17 \text{ km}^3$  per year. In fact, the majority of the Earth's crust was made in this way.

However, at depths below 60 km in the mantle, cold (solid) basaltic material converts to a form of rock called *eclogite*, which is much denser than shallow-mantle rocks because it contains garnet, a complex, aluminum-bearing silicate mineral. Large bodies of eclogite can sink through the upper mantle, which probably explains why the crust on Earth never gets thicker than about 60 km. Inside smaller terrestrial planets, like Mars, the pressures at a given depth are lower, so their crusts can extend farther down without converting to dense eclogite. On a hotter planet, like Venus, a thick basaltic crust would melt at its base rather than convert as a solid to eclogite.

Although our direct samples of the Earth's interior are limited to the crust and shallow mantle, we know from seismic tomography that broad regions with low seismic velocities extend to a depth of at least 400 km under oceanic ridges and other volcanic terrains. Magmas and rock-magma mixtures have low densities and low seismic velocities, so it seems reasonable that the basalt source region lies below about 400 km. When a hot silicate rock or low-density magmatic mush ascends from that great depth, it eventually separates into molten liquids (which erupt at volcanoes) and crystals (which stay behind in the mantle or form new crustal material).



**Figure 2.** Earthquakes trigger different kinds of diagnostic seismic waves that travel around the Earth and through its interior at 3 to 15 km per second. Compression (*P*) waves move almost twice as fast as shear (*S*) waves; they can also pass through the liquid outer core, which the *S* waves cannot.



**Figure 3.** This series of maps shows the state of the Earth's interior at different depths as determined by seismic tomography. In each, dark lines denote land masses, white lines show plate boundaries, and white circles mark the locations of "hot spots." The red regions have slower-than-average seismic velocities (4 percent in the upper panels, 0.5 percent in the lower ones) and are therefore hot. The blue regions have faster velocities and are therefore colder. In the 250-km map, derived from surface-wave data, notice the association of hot mantle with oceanic ridges and continental tectonic regions. The 380-km map, from shear-wave data, shows a large hot region in the central Pacific. Compression-wave velocities were used for the maps at depths of 1,000 and 2,500 km.



Thus, we have been able to identify three outer layers in the Earth: (1) the buoyant crust, containing low-density minerals dominated by quartz ( $\text{SiO}_2$ ) and metal-poor silicates called feldspars; (2) the uppermost mantle, containing minerals (primarily olivine and pyroxene) that are refractory (crystalline at high temperatures) and thus settle out of rising magma mushes; and (3) a "fertile" layer, below 400 km, that contains a large basaltic component and therefore abundant calcium and aluminum. This third layer is dense when cold, due to the garnet it contains, and buoyant when hot, because garnet and related minerals melt easily to form basalt.

Underneath all of this is the lower mantle. If the Earth has "cosmic" abundances of the elements, as deduced from their proportions in the Sun and primitive meteorites, then the lower mantle (with 70 percent of the mantle's mass) must be mainly silicon, magnesium, and oxygen. It probably also contains some iron, calcium, and aluminum. Although Ca and Al are well-represented in Earth's crust, the crust is too thin to yield Ca:Si or Al:Si ratios for the whole Earth as high as those found in the Sun, meteorites, and, by inference, the planets. Moreover, there is little calcium or aluminum in upper-mantle rocks (otherwise basalts could not rise through them en route to the surface), nor are they present in the core. So, by elimination, the bulk of Earth's calcium and

aluminum must reside in the lower mantle or in the mesosphere.

The lower mantle must also be richer in silicon than the layer above it. The reasoning behind this assumption is as follows: Primitive meteorites and the Sun have about one magnesium atom for every silicon atom. In the mantle, this 1:1 ratio would favor the formation of the mineral enstatite ( $\text{MgSiO}_3$ , a pyroxene) over forsterite ( $\text{Mg}_2\text{SiO}_4$ , an olivine). However, we know from its surface exposures that the upper mantle is olivine-rich and has a Mg:Si ratio of about two. Farther down, at the high pressures present in the lower mantle,  $\text{Mg}_2\text{SiO}_4$  decomposes to two new minerals. One is periclase ( $\text{MgO}$ ), which has the crystal structure of ordinary table salt, NaCl. The other is an ultrahigh-pressure form of enstatite (which, incidentally, has the same crystal structure as many of the new high-temperature superconductors). This enstatite variant propagates seismic waves at much higher velocities than periclase does and matches the seismic velocities we have observed for the lower mantle. Therefore, at great depths  $\text{MgSiO}_3$  would appear to be the most abundant mineral – and  $\text{MgO}$  largely absent. Some seismic evidence also indicates that the lower mantle has more iron (as FeO) than the upper mantle does; it may be similar to the mantles of the Moon and Mars, which we also suspect to be rich in FeO.

#### DIVIDING THE EARTH'S INTERIOR

The Earth's seismic properties have allowed geophysicists to distinguish rather distinct layering in its interior (Figure 5). In 1906, the British geologist Richard D. Oldham found that at a certain depth, compression or P waves slow sharply and S waves cannot penetrate further. It was the first evidence that the Earth has a liquid core. Only three years after Oldham's revelation, the Yugoslavian seismologist Andrija Mohorovičić discovered that the velocity of seismic waves takes a large jump about 60 km down. This Mohorovičić or "Moho" seismic discontinuity marks the crust-mantle boundary, where changes in rock chemistry and crystal structure occur. At the core-mantle boundary, averaging 2,890 km in depth, the composition of rock changes from silicate to metallic and its physical state changes from solid to liquid. This boundary is also known as the Gutenberg discontinuity, after Beno Gutenberg, who made the first accurate determination of its depth. Seismic discontinuities allow a further division of the Earth into inner core, outer core, lower mantle, transition region, upper mantle, and crust (Table 1). These regions are not necessarily all chemically distinct, nor can we assume that each of them is chemically homogeneous.

The inner core represents only 1.7 percent of the Earth's mass. It is solid, primarily the result of "pressure-freezing" (most liquids will solidify if the temperature is decreased or the pressure is increased). Probably the entire core was once molten, but over time it has lost enough heat for the inner core to solidify. It "floats" in the center of the outer core and is thus essentially decoupled from the mantle.

The outer core (30.8 percent of Earth's mass) is liquid, a result of its high temperature and the fact that iron alloys melt at lower temperatures than do common rocks. The viscosity of the outer core is very low, probably not much greater than water. We expect it to behave in general like other fluid parts of the Earth. Rapid motions of molten iron



**Figure 4.** The island of Hawaii consists entirely of outpourings from the Earth's mantle. The now-dormant volcano Mauna Kea, in the island's northern half, has become the site of numerous astronomical observatories. But Mauna Loa, to its south, is still quite active, especially along its southeast flank. The Hawaiian Islands are part of a long chain of peaks that formed as the Pacific lithospheric plate slowly moved northwest over a plume of upwelling mantle material. Hawaii, at the southeast end of the island chain, currently sits almost directly over the hot spot; the exact location is marked by a small submerged peak, Loihi, to the island's south. This photograph is a composite of two Landsat images.



in the core are responsible for the Earth's magnetic field and for some of the subtle jerkiness in our planet's rotation. The density of the outer core is slightly less dense than pure molten iron and requires about 10 percent of some lighter elements such as sulfur or oxygen, or both. These elements are considered likely because they are cosmically abundant and would readily dissolve in the hot metallic soup.

Just above the core is a 200- or 300-km-thick layer, called *D'*, that may differ chemically from the rest of the lower mantle lying above it. It may represent material that was once dissolved in the core, or dense material that sank through the mantle but was unable to sink into the core. The *D'* layer comprises about 3 percent the Earth's mass, or about 4 percent of the mantle.

Smaller seismic discontinuities occur at several depths in the mantle and halfway through the core. These are often attributed to phase transitions, but they may signify changes in composition. The two largest ones in the mantle, 400 and 650 km down, represent abrupt rearrangements of the atoms in the major mantle minerals. Large variations in seismic velocity have also been found from place to place. These "lateral variations" have been revealed by seismic tomography. In fact, the Earth's upper mantle exhibits as much variation horizontally as it does vertically (Figure 6).

#### WHERE ON EARTH IS THE CRUST?

Planets grow by colliding with other objects, an energetic process that results in melting or even vaporization. Most of the energy is deposited in the outer layers, except for the small number of truly giant impacts that are as likely to destroy the target object as add to its bulk; these may melt a large fraction of a planet. A global ocean of magma can segregate incoming material into solid and liquid fractions that float and refractory crystals and iron-rich melts that sink. Differentiation is akin to what takes place in a blast furnace or fat-rendering plant. By heating and boiling, the original material is reduced to frothy scum, dense dregs, and a "purified" liquid in between.

Planetary geologists have invoked such a global magma ocean to explain the Moon's anorthositic highlands (calcium- and aluminum-rich silicates that floated to the surface) and its "KREEP" basalts (which cooled from the final liquid dregs of a crystallizing magma ocean and thus became highly enriched in trace-elements). A similar process probably occurred inside the Earth, except the pressures were much higher, which caused dense garnet-bearing eclogite to form instead of buoyant anorthosite. In fact, high-grade anorthosite is fairly rare on Earth. Therefore, one key product of Earth's magma ocean did not float to the surface but sank from view. Had we not obtained actual anorthositic samples of the Moon, the magma-ocean concept might never have occurred to terrestrial geologists.

The crust of the Earth would be about 200 km thick if most of the low-density and easily melted material in the interior had separated out during Earth's formation. Yet the average terrestrial crust (20 km) is considerably thinner than the lunar crust (100 km) – even though the Moon has only 2 percent of the Earth's volume. Does this mean that the Earth did not have a magma ocean? Or has the crustal material mostly remained in or returned to the mantle?

The lunar crust is so thick and contains so much of the Moon's calcium and aluminum that it must have formed very

Region	Depth (km)	Percent of Earth's mass	Percent of mantle-crust mass
Continental crust	0–50	0.374	0.554
Oceanic crust	0–10	0.099	0.147
Upper mantle	10–400	10.3	15.3
Transition region	400–650	7.5	11.1
Lower mantle	650–2,890	49.2	72.9
Outer core	2,890–5,150	30.8	
Inner core	5,150–6,370	1.7	

Table 1. A summary of Earth's internal structure, as deduced from decades of probings with seismic techniques.

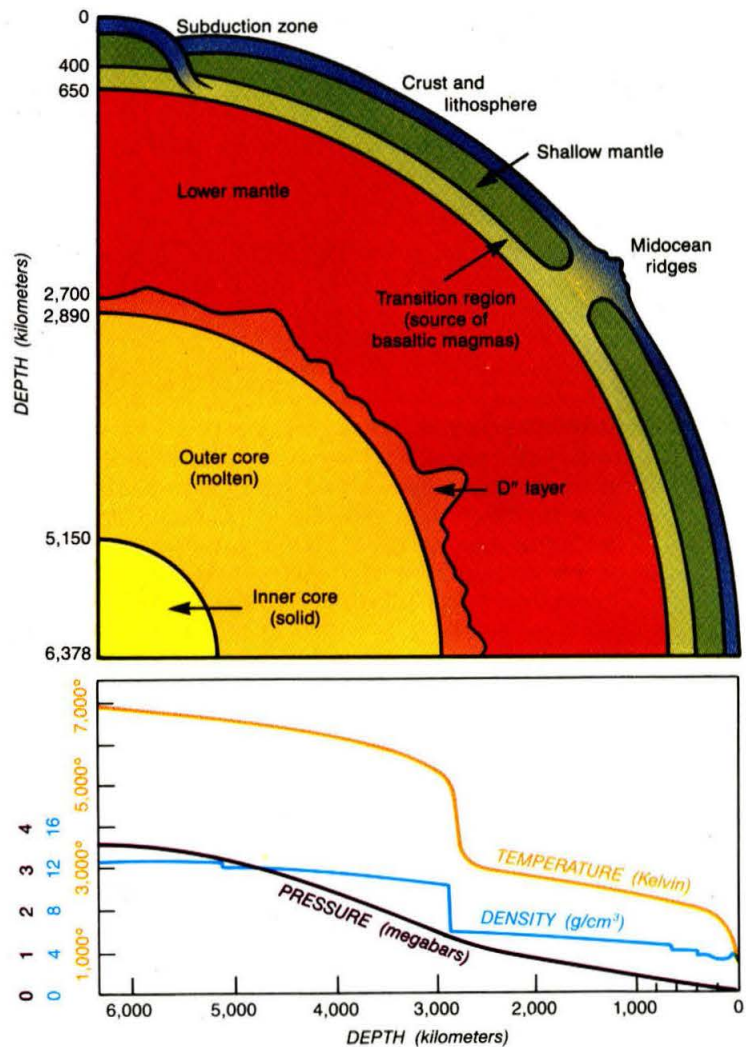


Figure 5. Early in its history, the Earth differentiated into a series of layers with distinct physical and perhaps compositional properties.

efficiently, for example, with its light crustal minerals rising directly to the top of a deep magma ocean. However, on a larger body like Earth, the pressures far down in a magma ocean are so great that buoyant minerals never form. Instead, dense crystals such as garnet and pyroxene soak up the calcium and aluminum. These, by and large, stay in the mantle and may even sink to the base of a magma ocean, thus limiting the crust's thickness. Even so, the high concentrations of some elements in the Earth's crust (Table 2) tell us that most of the mantle *must* have differentiated



**Table 2.** The abundance of various elements in the Earth's crust, as a percentage of their estimated abundance in the whole Earth.

Rubidium	(Rb)	68
Cesium	(Cs)	67
Thorium	(Th)	55
Barium	(Ba)	49
Uranium	(U)	47
Lanthanum	(La)	27
Strontium	(Sr)	21
Sodium	(Na)	13
Aluminum	(Al)	2.4
Calcium	(Ca)	0.9
Silicon	(Si)	0.7
Iron	(Fe)	0.07
Magnesium	(Mg)	0.06

either during accretion or shortly thereafter. These elements happen to be ones that are not easily incorporated into the high-pressure minerals that form at depth in a magma ocean. It is therefore unlikely that the Earth made its crust inefficiently but very likely that the missing crust resides somewhere in the mantle. The amount of crust now at the Earth's surface is much less than the *potential* crustal material and probably only a small fraction of the total volume of the crust that has been generated in  $4\frac{1}{2}$  billion years.

There is also a good reason why the Earth cannot have a thick "secondary" crust – that is, one formed by continental collision, mountain building, or the accumulation of volcanic materials. Wherever these processes cause the Earth's crust to thicken to more than about 60 km, the low-density crustal minerals convert to denser ones, causing the bottom of the crust to "fall off" or, technically, to delaminate. But even if delamination did not occur, the great pressure present below 60 km makes the seismic velocities there so high that a seismologist would call this deep-lying material part of the mantle, not the crust. In fact, "crust" is a physical concept, and its properties and thickness are derived from seismology. However, since erosion and volcanism supply us with many samples of the lower crust and shallow mantle, we know that the crust truly is compositionally distinct. As mentioned, it is calcium-, aluminum-, and silicon-rich compared to the

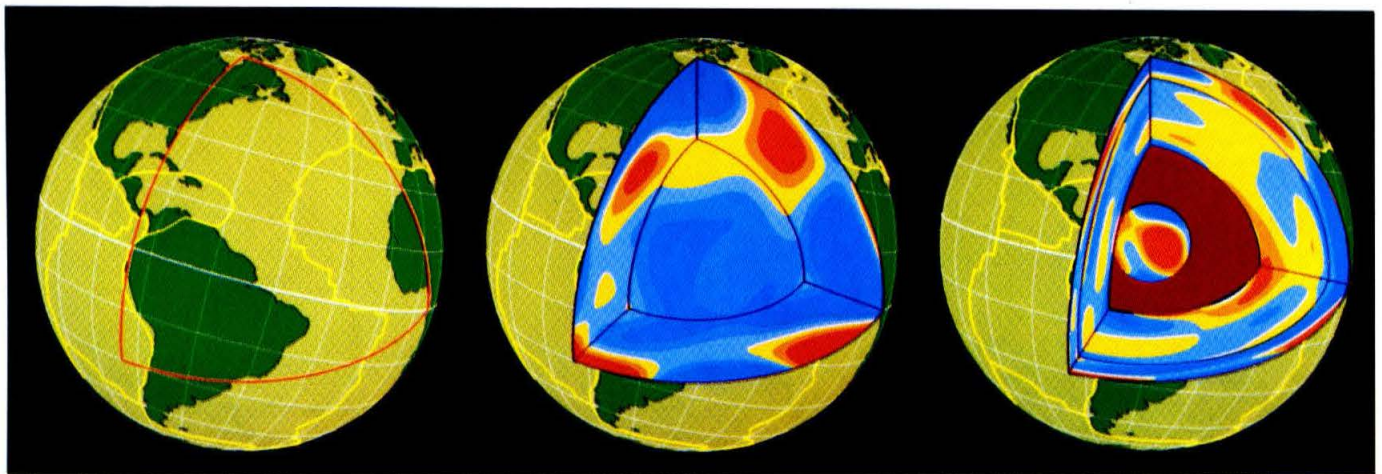
shallow mantle, so changes in physical properties at the crust-mantle boundary are accompanied by changes in chemistry as well.

### THE LITHOSPHERE

Although most of the crust and mantle are solid, we know from seismic velocities, the abundance of volcanoes, and the rise of temperature with depth in wells and mines that much of the outer part of the Earth is near or above its melting point. In fact, the coldest part of our planet is its surface. Since cold rocks deform slowly, we refer to this rigid outer shell as the *lithosphere* (the "rocky" or "strong" layer). On the Earth the lithosphere is not a single seamless shell, but rather a patchwork of rigid, snugly fitting *plates* that ride atop the mantle (Figures 7,8). These plates – eight large ones and about two dozen smaller ones – are moving with respect to one another, and their interactions are collectively called *plate tectonics*, a subject to be discussed later.

At depths between about 50 and 100 km, lithospheric rocks become hot and weaken enough structurally to behave as fluids – at least over geologic time. This portion of the upper mantle is called the *asthenosphere* (or "weak" layer), and it may be partially molten. Seismic velocities under young seafloor and tectonic regions are so low that some partial melting is required to depths as great as 400 km. Below the asthenosphere, the temperature continues to climb, but the solidifying effects of high pressure become dominant. So at still-greater depths the Earth again becomes strong and harder to deform. The region between the 400- and 650-km seismic discontinuities is called the *transition region* or *mesosphere* (for middle mantle), and the basalts that make up midocean ridges and new oceanic crust may be derived from this region.

Within the Earth's lithosphere, rocks are so cold and their viscosity so high that they support large loads and fail by brittle fracture rather than by deforming smoothly. New lithosphere forms at midocean ridges and thickens with time as it cools and moves away from the ridge, a process termed *seafloor spreading* (Figure 9). From the way it deflects under large submarine volcanoes and enters deep-sea trenches, we

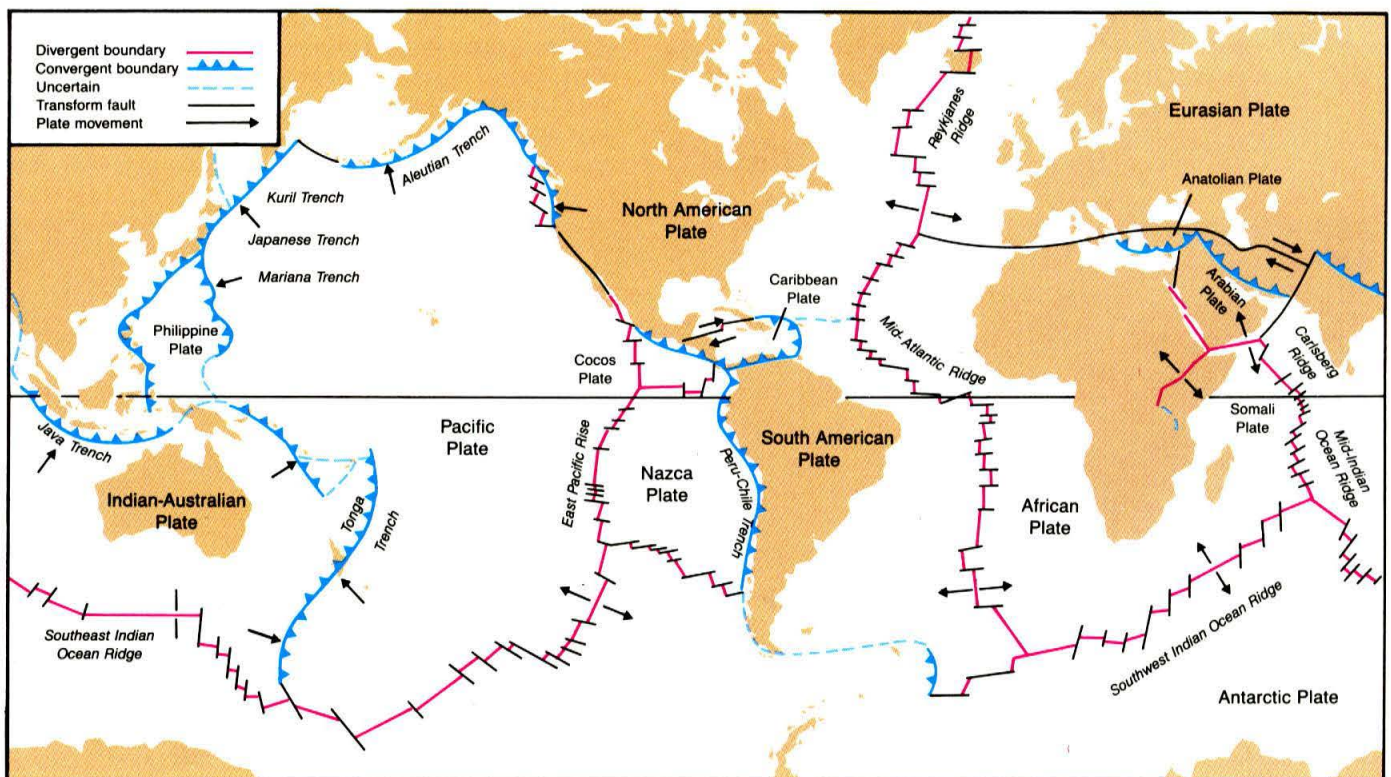


**Figure 6.** Earth's interior exhibits considerable variation with both depth and location. In these plots, red indicates the slowest seismic velocities and the hottest material, blue the fastest and coldest. The panel at left shows continents and plate boundaries (yellow lines) surrounding the Atlantic Ocean. The middle panel shows a slice into the underlying mantle to a depth of 550 km (with depth exaggerated by five times). The cutaway at right goes to a depth of 2,890 km, the mantle-core boundary; it also shows the solid inner core suspended within the outer core (red).





**Figure 7.** With the oceans emptied and the continents obscured, this map reveals a seafloor shaped by ceaseless geologic activity. This map uses radar altimetry from NASA's Seasat satellite of the ocean's surface, which reflects the underlying topography.



**Figure 8.** The Earth's major lithospheric plates are in motion with respect to one another. At divergent boundaries (such as midocean ridges) the plates move apart, only to collide and overlap at convergent boundaries (subduction zones). Plates slide past each other along transform faults, the most famous of which is the San Andreas fault that runs the length of California.



know that the oceanic lithosphere acts as an elastic plate whose thickness varies from zero at the midocean ridges to about 40 km under older seafloor. As it cools and moves aside, it also becomes denser and loses the high-temperature buoyancy it had initially; eventually, it tries to sink back into the interior.

Let us examine the formation and evolution of oceanic lithosphere a bit more closely. Portions of the Earth's mantle, especially the asthenosphere, behave like hot plastic and are in continuous, slow convective motion that brings heat from the interior to the surface. Upwelling mantle material partially melts in the asthenosphere, where a segregation takes place; denser refractory crystals are left behind when the lighter, easily melted material erupts upward. This buoyant, erupting melt creates the lithosphere's topmost layer, the oceanic crust, which is basaltic and averages about 6 km in thickness.

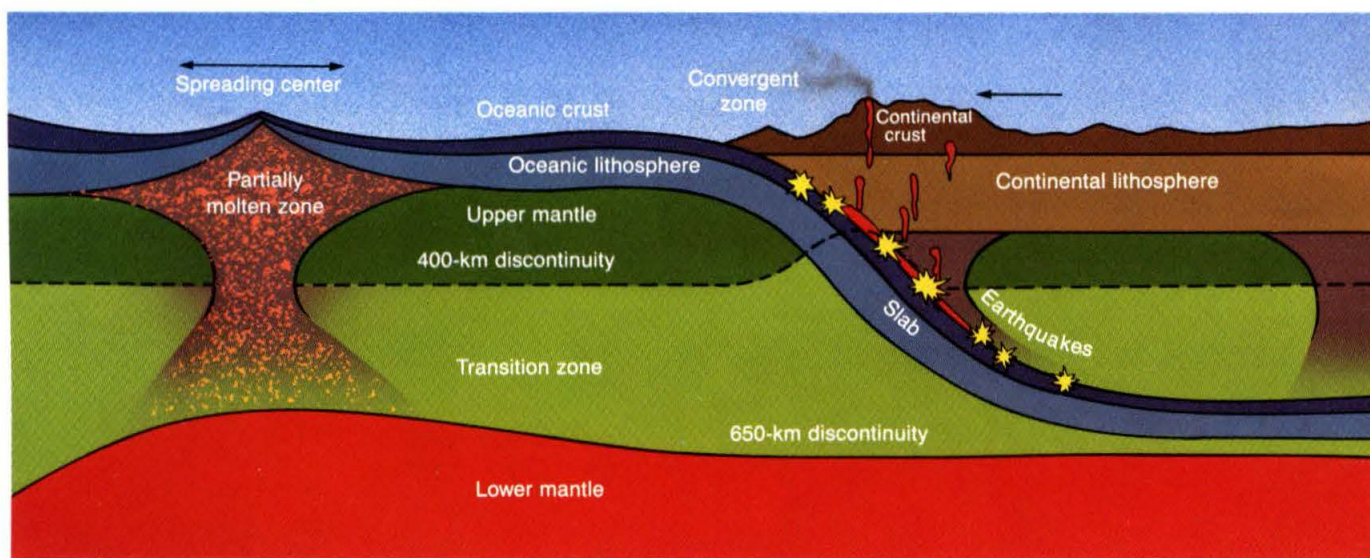
The lithospheric layer beneath this is essentially normal mantle material that has lost its basaltic component. The basalt is missing either because it rose as a melt to the crust or, at depths of roughly 60 km, it reverted to dense eclogite and sank as large blobs through the upper mantle and into the mesosphere. However, melting of eclogite-rich mesosphere material restores its buoyancy, and the resulting magma mush will rise toward the surface. If it has a clear upward path, as occurs along the midocean ridges, it will erupt as basalt onto the seafloor. The magma does not always ascend vertically, however, and may first migrate laterally through the mantle for great distances. Elsewhere its path may be blocked completely, so it pools on the underside of previously formed lithosphere. We are ignorant of the composition of the lower oceanic lithosphere, but it is probably a mixture of basalt (which ultimately converts to eclogite as the lithosphere cools and thickens) and refractory crystals. Substantial amounts of eclogite in the older lithosphere would help explain why it eventually sinks back into the interior.

In contrast, the continental lithosphere is about 150 km

thick, and its crustal and upper-mantle components are both buoyant relative to the normal mantle below. Continents therefore float around as icebergs and do not directly participate in the deeper circulation currents of mantle convection. But lateral movement in the mantle can and does move these lithospheric icebergs around, and once they come to rest they can insulate the underlying mantle and cause it to warm. In the course of this *continental drift*, continents can override the thinner oceanic lithosphere along *subduction zones* – linear or arcuate features characterized by deep oceanic trenches and large volcanic cones. If the oceanic lithosphere is still young and thus hot, it tends to slide under the continent at a shallow angle; older, thicker lithosphere is denser and tends to dive steeply into the mantle.

On Earth and elsewhere, the lithosphere is an important element in planetary dynamics. If it gets too cold or too thick, it can shut off the access of hot magmas to the surface or become too hard to break and descend (subduct). If its proportion of light minerals is too great, it will stay buoyant and will not sink back into the mantle. If there are too many plates or if they are moving rapidly, they again may not become dense enough to subduct. Thus, there are a variety of ways to “choke up” the surface. In the extreme, a lithosphere may get too thick to break anywhere, creating one uninterrupted plate that can slide around as a unit on the underlying mantle. On such a “one-plate planet,” a huge meteoritic impact or the mass of a large new volcano could alter the planet's moment of inertia enough to make the whole outer shell rotate with respect to the spin axis.

Several mechanisms can fragment a lithosphere. Hot mantle upwellings can both heat and deform it. Diverging mantle currents below can create extensional stresses on its base. A lithosphere moving over an ellipsoid-shaped (rotating) planet will experience large stresses due to the changing contour of the surface. Tidal despinning of a planet is a related method for generating large stresses in the surface layer and a global fracture pattern. If the lithosphere becomes too dense it may sag and break. Several subduction



**Figure 9.** The basalt that forms oceanic crust does not come from immediately below the lithosphere but from a much-deeper transition zone in the mantle. As it rises, this material decompresses and may become partially molten; finally it erupts at a spreading center. As the new oceanic crust moves away from its formation site, it cools and thickens, eventually becoming dense enough to plunge back into the mantle. This subduction occurs dramatically along zones of convergence marked by deep trenches, frequent earthquakes, and active volcanism.



zones currently exist entirely under the Earth's oceans. But it is not clear if they are the result of an instability of the oceanic lithosphere, or if subduction started at the edge of a continent and later migrated toward the ocean. We do know that both oceanic ridges and island arcs can migrate relative to the underlying mantle and the spin axis.

The spin axis of a planet is controlled by the distribution of masses on the surface and in the interior. By analogy, the rotation of a spinning top is controlled by its shape, and its spin axis will change if bits of clay are attached to the surface. The physics of planetary reorientation is the same. If a large impact or a new volcano redistributes the mass, the planet will reorient itself relative to the spin axis so that the mass excesses lie closer to the equator. This shift is termed *true polar wander*. Both Mars and the Moon have apparently reoriented themselves to accommodate the effects of impacts or volcanoes.

On the Earth at present, true polar wander is very slight and results mainly from the rearrangement of mass due to melting glaciers. Polar motion in times past, as evidenced by magnetically aligned crystals in ancient rocks, is usually considered to be *apparent polar wander*, since we know that the continents have been drifting relative to the magnetic pole. In addition, however, Earth's rotation axis has apparently moved about 8° in the past 60 million years and 20° in the past 200 million – a period of time when the configuration of continents and subductions zones was also changing dramatically. Our extraterrestrial experiences now tell us that major shifts of the Earth's lithospheric shell relative to its spin axis might have followed convective rearrangement of mass in the interior, plate subduction, or the build-up of heat beneath large continents. These mass adjustments might be responsible for some major events in the geologic record, such as the breakup of supercontinents discussed in the next section.

Although plate tectonics, or at least “seafloor” spreading, may exist on Venus (see Chapter 7), that planet does not have the more obvious manifestations of terrestrial-style plate tectonics such as long linear ridges, subduction zones, and deep trenches. Planetary geologists have recently found evidence in Venus' equatorial highlands for some crustal extension, with short ridge segments and fracture zones, but there is as yet no hint of the subduction process. Since Venus and Earth are so similar in size, why do they differ so much in tectonic style? One reason is that Venus spins much more slowly and therefore has a much smaller tidal bulge. Its shape is thus nearly spherical, so its lithosphere experiences no large stresses while moving around. Another reason is that the surface of Venus is much hotter than the Earth's, which makes its lithosphere thinner and more buoyant (particularly if the two planets' crusts are compositionally similar). At present we do not know which of these explanations is correct, or whether other factors await discovery.

## PLATE TECTONICS

Planets have various options for relieving themselves of their internal heat. The Earth chooses the plate-tectonic option, and most of our planet's interior heat is removed by this mechanism. Plate tectonics begins by the creation of new crust and upper mantle at long, globe-encircling cracks – the midocean ridges. While Arthur Holmes suggested that the

oceans were a source of new crustal material as long ago as the 1920s, it was not until the early 1960s that Harry Hess (and later Robert Dietz) refined the scenario of a dynamic, self-renewing seafloor and focused attention on the midocean ridges and deep-sea troughs. The associated volcanism occurs mostly underwater, but the ridges can be traced around the world by their bathymetry and their seismic activity.

Newly formed lithosphere cools and contracts as it moves away from a ridge. Consequently, the ocean depth above it increases in a smooth and characteristic way as a function of distance from the ridge and, therefore, of age. The oceanic lithosphere also thickens with age and eventually becomes denser than the mantle material below; in response, it sinks back into the mantle at subduction zones.

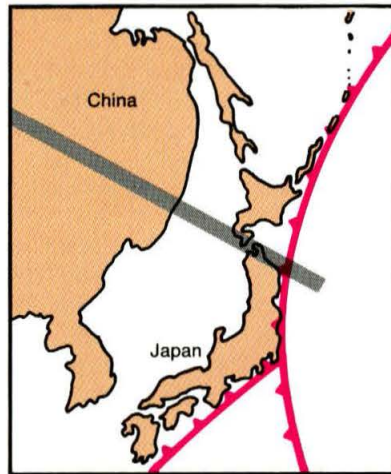
Most of the ocean floor is less than 90 million years old, and nowhere is it older than 200 million years. It takes about 200 million years for the oceanic lithosphere and shallow mantle to cool to a depth of about 100 km, and when this is inserted back into the hot mantle it becomes, in effect, an ice cube in a warm drink. Subduction is the main mechanism by which mantle deeper than 100 km cools. Earthquakes have been recorded at depths as great as 670 km, and geophysicists believe the cold oceanic plates, or slabs, can sink this far into the mantle (Figure 10).

The Earth is apparently unique among the known worlds in its use of deep subduction as a cooling mechanism, and this can only occur if the lithosphere gets cold enough to cause it to become unstably dense and sink. On a planet with a thicker crust, a hotter surface, or a colder interior, the lithosphere may be permanently buoyant. In fact, on Earth the continents *are* permanently buoyant, a combination of thick low-density crust capping a buoyant upper-mantle “root” extending down to about 150 km.

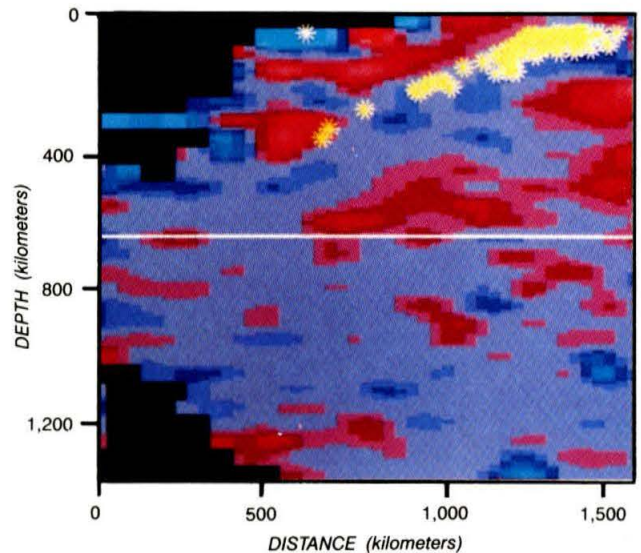
Smaller planets cool more rapidly than large ones, have lower gravity, and experience less vigorous internal convection. Therefore, a lithosphere of a given thickness would be harder to break up on planet smaller than Earth. The Moon (with 1 percent of Earth's mass) and probably Mars (11 percent) are single-plate planets. Their interiors can convect heat outward, but they are never exposed to the cooling effect of subducting lithosphere because their outer layers behave as more-or-less rigid shells. Except for isolated volcanoes, they must lose their internal heat by conduction. Mantle upwellings can focus heat on one portion of the shell, weaken and thin it, and permit magmas to erupt onto the surface. This situation occurs on Earth as well; variously called midplate volcanism, hot spots, and plumes, it accounts for about 10 percent of the heat flow from the terrestrial interior (Figure 11).

The Earth actually exhibits at least three tectonic styles. The oceanic lithosphere recycles itself. The continents are buoyant; they may break up and reassemble, but they remain at the surface. A third characteristic is the way continents affect and are affected by the underlying mantle and adjacent plates. They are maintained against erosion – rejuvenated, in a sense – by compression and uplifting (mountain building) at their boundaries with other plates, by the sweeping up of island arcs at their leading edges, and by eruptions of basalt onto, into, or under the continental mass. Heat escapes from below a continent mostly by conduction, a relatively slow process. Therefore, the underlying mantle can heat up





**Figure 10.** Seismic tomography has been used to create a cross-section of the subduction zone of northern Japan. In the red regions, compression waves travel at relatively slow velocities, indicating hot regions. Earthquakes (yellow stars) fall near the top of a cold (blue) slab of descending lithosphere.



enough to melt partially or, at least, to offset the cooling action of subducting oceanic lithosphere.

Since material flows from hot to cool parts of a convecting system continents will tend to drift away from hot mantle zones and come to rest over cool ones. When viewed from Africa, the continents are drifting away from each other at rates of some 5 to 10 cm per year. When this motion is traced back in time, we find that about 180 million years ago the continents were assembled into a supercontinent called Pangea (Figure 12). Moreover, for at least several hundred million years prior to that, the southern continents (Africa, South America, Australia, and Antarctica) plus India were a single assemblage, Gondwana. About 360 million years ago, Gondwana was centered on the South Pole, but it moved toward the equator just prior to its breakup. Initially, the continents' separation was rapid, but it slowed as the distances between them increased. As the continents moved apart, the Atlantic Ocean opened up and the Pacific Ocean shrank. Part of the Pacific lithospheric plate disappeared beneath the continental plates surrounding it.

Most of the continents are now sitting on or moving toward cold parts of the mantle. The exception is Africa, which was the core of Pangea. As they move around, the continents encounter oceanic lithosphere and force it to subduct into the mantle. Many active subduction zones are currently at the leading edges of continents. Perhaps all such zones formed along continental margins, after which some of them migrated to their present midocean locations.

There is another conceivable type of plate tectonics. If a large temperature difference does not exist between the surface and the interior, or if plate generation is very rapid, or if the crust-lithosphere system is completely buoyant – then deep subduction cannot happen. Consequently, the plates must remain near the surface, and their interactions will result in “pack-ice” underthrusting (much the way ice flows behave in the polar oceans). The convergence zones will be diffuse, elevated jumbles characterized by deformation, plate thickening, shallow underthrusting, and lithospheric doubling. Venus and the early Earth may have experienced this tectonic style, for we still see evidence of it in western North America, parts of western South America, and Tibet.



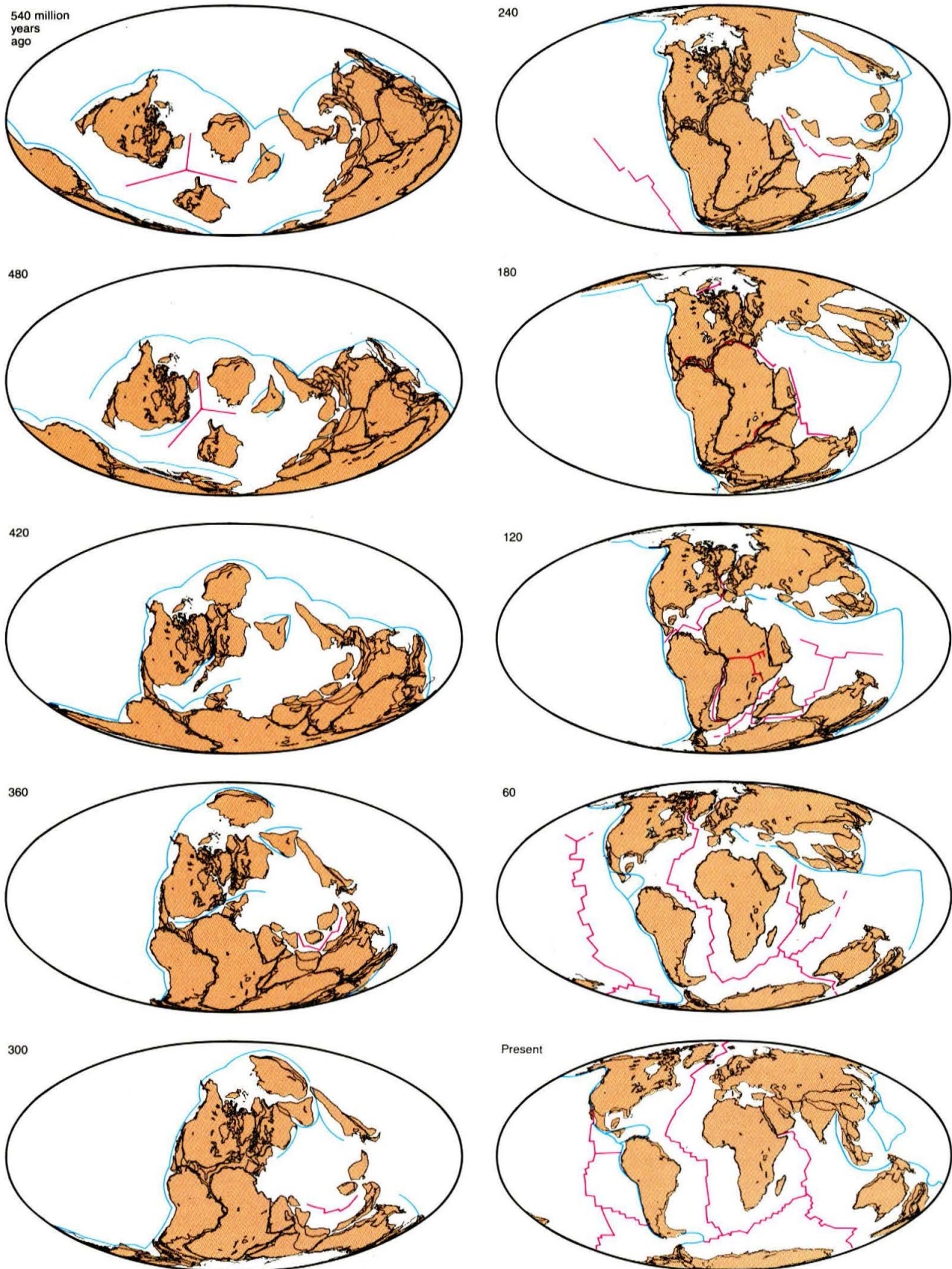
**Figure 11.** Iceland sits in the Atlantic Ocean atop both a midocean ridge and a stationary hot spot in the upper mantle. As a result, the entire island is alive with volcanic and tectonic activity. Underground water can become superheated, reaching the surface through fractures. Shown here is the Namaskard geothermal area in northeast Iceland.

## THE EARTH'S GEOID

On an entirely fluid planet the shape of its surface – the *geoid* – is not controlled solely by rotation. Concentrations of mass in the interior (actually, pockets of anomalously high density) attract the fluid, cause it to pool above them, and make the regional surface stand high. The geoid is usually defined with respect to the perfect ellipsoid that the planet would assume if its interior were completely fluid, with density changing only with depth. The result on a real planet is a global pattern of broad undulations, with heights of some hundreds of meters and a variety of wavelengths.

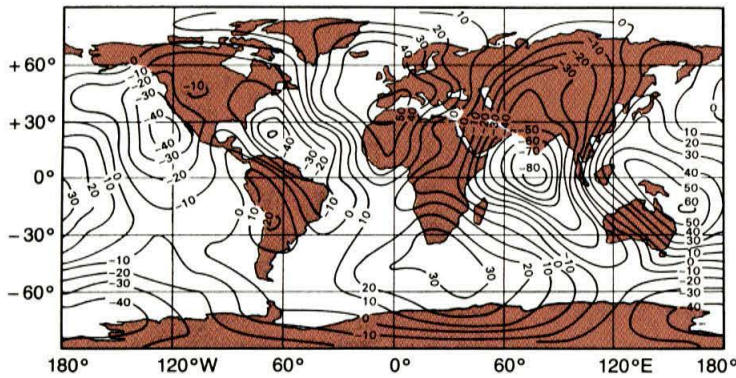
On Earth, the surface of the ocean approximates the geoid, but a more accurate figure for the entire planet has been obtained by tracking the motion of low-altitude satellites (Figure 13). While these geoid, or gravity, data cannot identify subsurface structures unambiguously, they can be used to calculate the contribution from isostatically





**Figure 12.** The Earth's face has changed dramatically in the last half billion years, as shown here in 60-million-year intervals. Note the assembly of Gondwana at the south pole prior to its incorporation into the supercontinent Pangea. Pangea moved northward across the equator over 150 million years; its eventual breakup created the Atlantic Ocean and greatly diminished the extent of the Pacific Ocean.





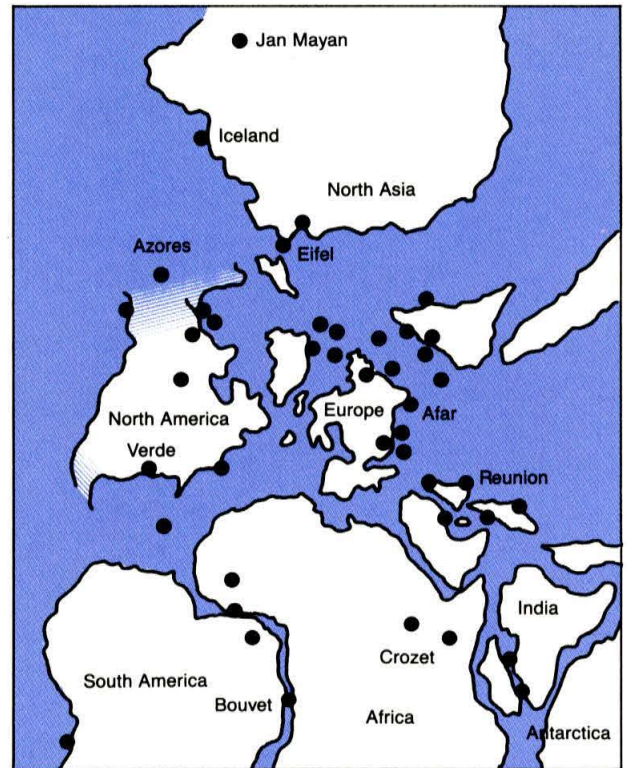
**Figure 13.** After allowing for the Earth's polar flattening (about 22 km, or 0.34 percent), geophysicists have analyzed satellite-tracking data to reveal that our planet displays residual geoid highs and lows, given here in meters with respect to the ideal spheroid. The highs correspond to subsurface concentrations of dense rocks, the lows to accumulations of less-dense rocks probably within the mantle.

compensated continents, slabs, and density variations in the lower mantle. (A continent is considered to be in a state of *isostasy* if equilibrium exists between gravity's downward pull on the mass sitting above sea level and the upward push of the mantle on the continent's low-density "root." Icebergs, in a sense, float isostatically in sea water.)

At very long wavelengths, there are equatorial geoid highs centered on the Pacific Ocean and Africa. Geoid lows occur in a polar band extending through North America, Brazil, Antarctica, Austria, and Asia. Brad Hager, Robert Clayton, and Adam Dziewonski have shown that this pattern correlates with the seismic velocity distribution in the lower mantle, as expected. The long-wavelength geoid highs arise from upwellings of hot mantle material that deform the core-mantle boundary and the Earth's surface upward. At the same time, the hot upwelling mantle is expected to be buoyant and thus relatively low in density, which seismic waves travel through more slowly than in cold material elsewhere. Except for Africa the continents are in or near geoid lows. We think they migrated into these regions as they moved away from Africa after the destruction of Pangea.

The major geoid highs of moderate wavelength are associated with subduction zones stretching from New Guinea to Tonga and along the Peru-Chile coastline. These highs, centered on the equator, undoubtedly contribute to the moment of inertia that controls the orientation of the Earth's spin axis relative to its mantle. At shorter wavelengths subduction zones show up as geoid highs, or mass excesses. This is expected as long as the descending slabs are cold, dense, and supported from below by a strong or dense lower mantle.

From Figure 13, it is apparent that Earth's present-day expressions of tectonism correlate poorly with its geoid. However, there is *good* correlation with the continental and subduction-zone configurations of the past. For example, the



**Figure 14.** The locations of hot spots about 350 million years ago. Today hot spots are primarily in the Atlantic and Indian oceans and under Africa, but most of them were originally underneath Pangea. They may have come about because the great mass of continental material kept heat from leaving the mantle, causing a buildup of thermal energy that has taken hundreds of millions of years to be released.

geoid high centered over Africa has about the shape and size of Pangea, and geoid lows correspond roughly with where regions of subduction should have existed prior to extensive opening of the Atlantic Ocean. This is an excellent demonstration of the time-scales on which planetary processes operate – the heat trapped under the supercontinent of Pangea more than 100 million years ago (Figure 14) continues to escape from the mantle today. The still-hot mantle has thus elevated the continent of Africa; it represents a geoid high.

We have been studying the Earth's surface since our arrival here, but only within the past several decades have we come to appreciate the internal turmoil that continuously shapes the landscapes around us. We were learning about the roles of continental drift and plate tectonics on the Earth at the same time we realized that every other world has a unique style of operation. As far as we know the Earth is the only planet that has active plate tectonics, oceans, and life. One wonders if these facts are interrelated.